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Closing a sediment budget for a reconfigured reach of the Provo River, Utah, United States

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[1] We quantified all components of a fluvial sediment budget for a discrete flood on an aggrading gravel bed river. Bed load transport rates were measured at the upstream and downstream ends of a 4 km study area on the Provo River, Utah, during a dam-controlled flood. We also collected high-resolution measurements of channel topography before and after the controlled flood for the entire reach. Topographic uncertainty in the digital elevation models (DEM) was characterized using a spatially variable approach. The net sediment flux provided unambiguous indication of storage. Sediment input to the reach (319 m^3) exceeded output (32 m^3), producing a net accumulation of approximately 290 m^3. The difference between the scour and fill was also positive (470 m^3), but uncertainty in the topographic differencing was larger than the observed net storage. Thus, the budget would have been indeterminate if based on morphologic data alone. Although topographic differencing was not sufficiently accurate to indicate net storage, it was able to demonstrate that internal erosion was a larger sediment source than the net sediment flux. The magnitude of total erosion (1454 m^3) and deposition (1926 m^3) was considerably larger than net change in storage, showing that internal sources and sinks were the dominant driver of channel change. The findings provide guidance for the development of sediment budgets in settings in which one must choose between a morphological approach and the direct measurement of sediment flux.

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1. Introduction

[2] Sediment budgets are a fundamental tool in geomorphology, used across the discipline in theoretical and applied studies [*Reid and Dunne*, 2003]. A fluvial sediment budget provides the context needed to evaluate channel response to changes in flow or sediment supply [e.g., *Trimble*, 1983; *Wathen and Hoey*, 1998; *Grams and Schmidt*, 2005]. A sediment budget balances sediment input (I) and sediment export (E), against sediment storage (ΔS),

$$I - E = \Delta S. \quad (1)$$

Numerous studies have focused on quantifying either the flux side [*Singer and Dunne*, 2004; *Vericat and Batalla*, 2006] or the storage side of (1) [*Lane et al.*, 1995; *Martin and Church*, 1995; *Ashmore and Church*, 1998; *Ham and Church*, 2000; *Brasington et al.*, 2003; *Surian et al.*, 2009]. Few studies have computed both sides of the budget. In the absence of closing a budget, the unmeasured components of the budget cannot be separated from the errors associated with the measured terms in the budget [*Kondolf and Matthews*, 1991]. Closure of the budget, i.e., independently calculating the left and right sides of (1) and determining if the two quantities match, provides a rigorous means by which the accuracy and precision of the budget can be evaluated. Spatial partitioning of the right side of (1), i.e., determining the amount of change in sediment storage in different parts of the channel and/or floodplain, provides even more insight into how channels adjust to longitudinal changes in sediment transport.

[3] One of the persistent problems with developing sediment budgets is that measurement error is typically large. Both transport and storage sides of the budget often involve the small difference between two large and uncertain numbers, such that even the sign of either side of (1) is uncertain. Sediment transport estimated from either formulas [*Gomez and Church*, 1989; *Martin*, 2003] or direct measurement [*Ham and Church*, 2000; *Wilcock*, 2001] may not be sufficiently accurate to determine the sign of the net flux. Topographic monitoring may not be sufficient to determine the sign of ΔS, even with recent advances in techniques for measurement and analysis [*Heritage and Hetherington*, 2007; *Wheaton et al.*, 2010; *Milan et al.*, 2011]. The result of measurement uncertainty in either flux or topographic change is that sediment budgets may be indeterminate, in

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the sense that one cannot explicitly demonstrate that aggradation or degradation has occurred [*Grams and Schmidt*, 2005]. In some cases, budgets are indeterminate even where measurement programs are extensive [*Topping et al.*, 2000].

[4] The ability to calculate a sediment budget with a definitive balance is influenced by the spatial and temporal scale of the analysis. For example, budgets may be developed over sufficiently long reaches such that there is a significant difference in fluxes at the upstream and downstream boundaries or over sufficiently short temporal scales so one can accurately relate topographic measurements to a discrete flow event. Even under these circumstances, however, challenges remain. Detecting the changes in storage for short time spans may be difficult because there may have been little net topographic change during the period for which the budget is calculated. It may be difficult to extrapolate the budget to longer time scales because a longer time span introduces more uncertainty about the stability of sediment transport relations. Additionally, investigations spanning multiple years, or decades, are often limited by the lack of historic data. Calculation of budgets over short spatial scales provides the advantage that changes in storage can be measured with relative ease. However, when budgets are calculated for a short reach of river, there may not be a significant difference between the measured influx and efflux of sediment. Conversely, larger spatial scales provide the advantage that there may be a more substantial difference between influx and efflux. Yet, changes in storage are more difficult to comprehensively measure over longer reaches.

[5] Here we present a sediment budget for a reconfigured 4 km segment of the middle Provo River, near Heber City, Utah, USA (Figure 1) for a single flood that lasted approximately 3 weeks. We highlight the challenges and uncertainties associated with construction and closure of a sediment budget in an unusually well-constrained situation—a discrete flood on a relatively short segment of a gravel bed river. The study area had been reconfigured approximately 3 years earlier, and qualitative evidence indicated that the channel was accumulating gravel, primarily in point bars. Preliminary measurements of transport rates at the upstream and downstream boundaries of the study area suggested that sediment influx exceeded efflux by an order of magnitude [*Olsen*, 2006]. In constructing a sediment budget, we sought to (1) confirm whether or not aggradation was occurring, (2) understand the magnitude of difference between upstream sediment delivery and downstream sediment export, and (3) evaluate whether the observed channel changes could be attributed to sediment accumulation in the reach. By quantifying both sides of (1), this study provides guidance and insights into the merits, and demerits, of using either measurements of flux or morphologic change in understanding geomorphic systems.

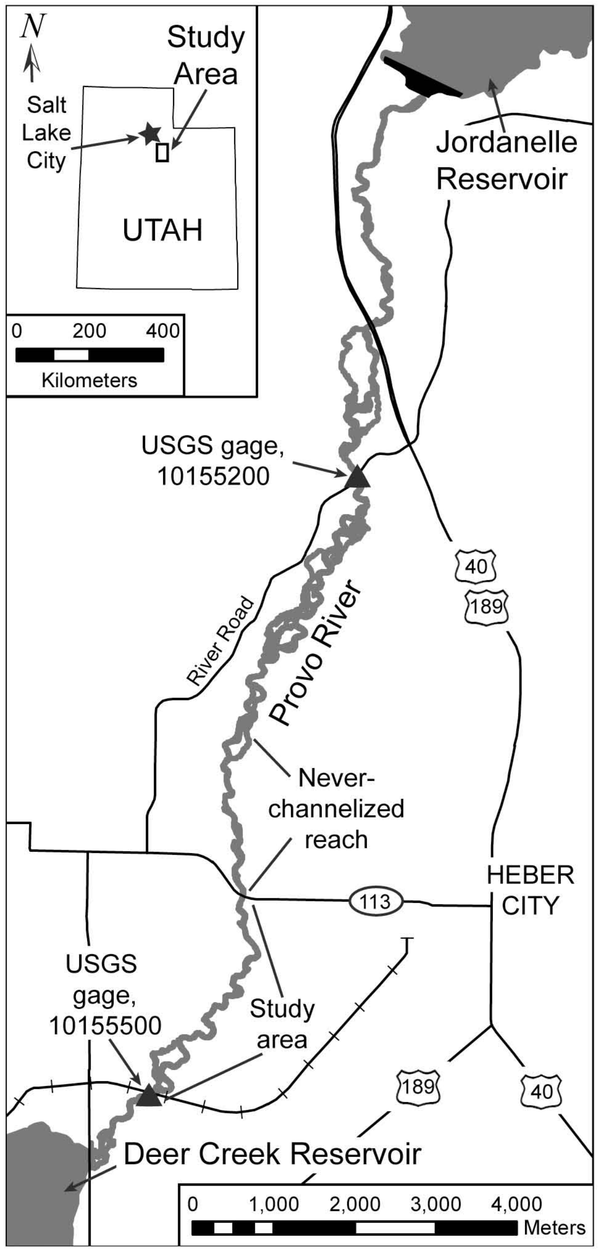

Figure 1. The middle Provo River, located in Heber Valley, Utah. The river flows approximately 19 km from the outlet of Jordanelle Dam (top right) to Deer Creek Reservoir (bottom left). The entire channel was reconfigured as part of the Provo River Restoration Project, with the exception of the Never Channelized Reach (NCR). Today, the NCR provides a local source of sediment to the study area.

2. Study Area

2.1. History of Flow Manipulation on the Provo River

[6] The Provo River flows from its headwaters in the Uinta Mountains in northern Utah to its outlet in Utah Lake. The river system is subject to large-scale flow manipulation and augmentation, primarily caused by two water resource development projects: the Provo River Project (PRP) and the Central Utah Project (CUP). As part of the PRP, trans-basin diversions from the Weber and Duchesne Rivers into the Provo River were constructed in 1930 and 1952, respectively. These diversions nearly doubled the magnitude of peak flows on the Provo River and substantially increased base flows. Deer Creek Dam was constructed at the downstream end of Heber Valley to provide reservoir storage for the augmented flow (Figure 1). Deer Creek Reservoir was filled soon after completion of the dam in 1941. To accommodate the additional flow and to protect adjacent lands from flooding, the Provo River in Heber Valley was

straightened, enlarged, and confined between dikes along the entire length of the valley between 1960 and 1965. The only exception was a 2.5 km reach, which we refer to as the Never Channelized Reach (NCR; see Figure 1). During the period between 1965 and 1994, upstream bed incision caused gravel to accumulate in the NCR.

[7] Jordanelle Dam, at the upstream end of the Heber Valley, was completed in 1993 as part of the CUP. Operations of Jordanelle Dam reduced the magnitude of peak floods by 25% from those of the postflow-augmentation period. Additionally, trans-basin diversions maintain summer flows that are much higher than natural base flows. Within the study area, streamflow has been measured since 1938 at U.S. Geological Survey gauging station 10155500 (Provo River near Charleston). There are no significant tributaries.

2.2. Provo River Restoration Project

[8] The Provo River Restoration Project (PRRP) involved reconfiguration of 16 km of the middle Provo River to restore elements of the prechannelization ecosystem that can be maintained by the regulated flow regime provided by Jordanelle Dam. Twelve km of the PRRP are upstream from the NCR and 4 km are downstream from the NCR. Project construction began in 1999 and consisted of removing dikes, creating a wandering, gravel bed channel, reconnecting the river to existing remnants of historic secondary channels, and constructing small side channels to recreate natural aquatic features and wetlands. The reconstructed channel morphology is intended to maximize diversity of habitat conditions and establish a complex template on which ecosystem processes will thrive [*Utah Reclamation*, 1997].

[9] Jordanelle Dam eliminated the sediment supply once delivered to the Heber Valley; thus, the upstream 12 km of the PRRP has no sediment input. Previous transport observations [*Olsen*, 2006] indicated that the NCR is now a source of gravel for the reconfigured 4 km segment downstream. Air photo observations in this segment indicated that point bars had grown since completion of channel reconfiguration in 2004 (Figure 2), suggesting a trend of sediment accumulation.

[10] This study focuses on the 4 km segment immediately downstream from the NCR where qualitative observations suggested sediment is actively accumulating. The substantial influx of gravel into the study area (Figure 3) has the potential to augment channel dynamics, and perhaps aid in achieving restoration goals, but these channel changes also have the potential to be detrimental to the original restoration objectives. Thus, it is useful to determine the sediment balance in order to better understand the impact of sediment influx on channel morphology and dynamics. The PRRP upstream from the NCR has no sediment supply and gravel augmentation will be considered as a tool for promoting channel dynamics. Studies of the segment downstream from the NCR will inform plans for gravel augmentation plans upstream.

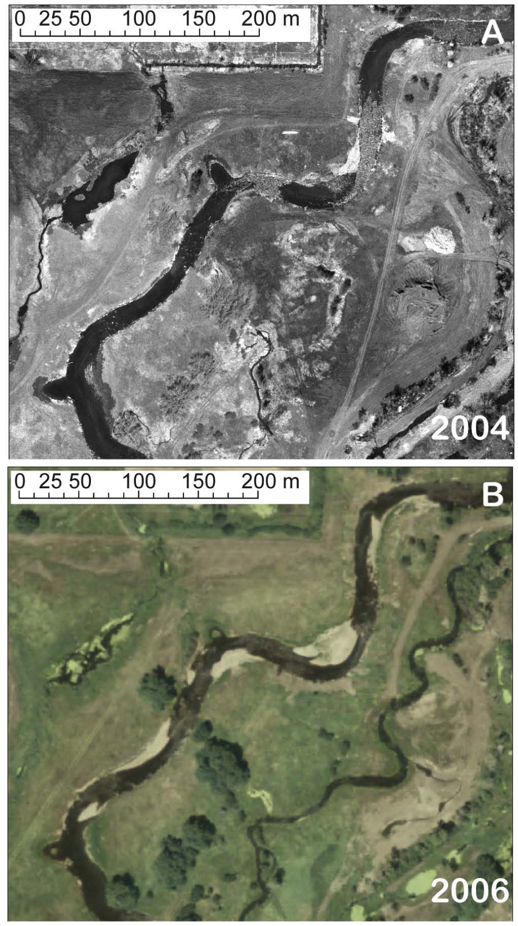

Figure 2. Aerial photos of reach 4 taken in (a) 2004 and (b) 2006. The 2004 image (Figure 2a) was taken shortly after reconfiguration of this reach. Figure 2b depicts point bars that grew during floods in 2005 and 2006. The location of the reach is shown in Figure 3.

3. Methods

[11] We quantified both net flux and change in storage, quantified their uncertainty, and evaluated the mass balance with respect to the uncertainty in all terms. Here we divide our discussion of methods into the quantification of bed load flux (section 3.1) and estimation of change in storage through measurement of topographic change (section 3.2).

3.1. Determining Bed Load Flux

3.1.1. Bed Load Transport Measurements

[12] In spring 2009 we worked with the U.S. Bureau of Reclamation (USBR) and the Central Utah Water Conservancy District (CUWCD) to design a controlled flood that allowed for an effective bed load sampling program. On both rising and falling limbs, discharge was changed in intervals of approximately 5.7 m^3 s^{-1} each day and then held steady for at least 8 h (Figure 4), allowing us to collect bed load measurements at the same constant flow rate at two sites. We collected transport samples at discharges ranging from 22.7 to 53.5 m^3 s^{-1}. The peak of the 2009 flood had a recurrence interval of 4 years for the 17-year record following closure of Jordanelle Dam.

[13] We established bed load transport measurement sites at the upstream and downstream boundaries of our study area, which we refer to as Midway and Charleston, respectively. At each bed load sampling site (Figure 3), we used a raft-based sampling platform [*Graham Matthews*

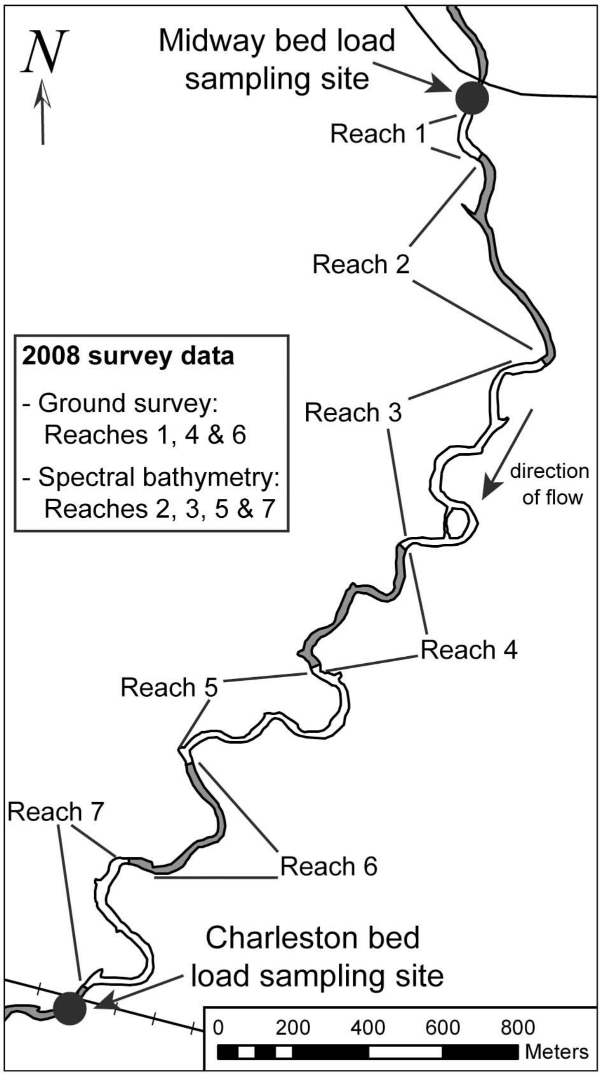

Figure 3. The study area: The lower 4 km of the Provo River Restoration Project (PRRP). A local sediment source, provided by the Never Channelized Reach (NCR), is located immediately upstream of the Midway sampling site (Figure 1). In 2009 we measured bed load transport at the upstream (Midway) and downstream (Charleston) sampling sites. Morphologic change associated with the 2009 flood was measured in each of the study reaches.

and Associates, 2010] and a Toutle River 2 (TR-2) bed load sampler [*Childers*, 1999]. The TR-2 sampler is well suited for measurements of the large grain sizes in transport during Provo River floods and the sampler has been used successfully on other large rivers [*Gaeuman et al.*, 2009; *Wallick et al.*, 2009; *Erwin et al.*, 2011]. We used a modified version of the equal width interval [*Edwards and Glysson*, 1988] sampling method; one sample consisted of a single pass across the channel, during which data were collected at 8–10 points along the cross section. The sampler remained on the bed for 5 min at each sampling station. We sieved and weighed all samples in 1/2-ϕ size classes.

[14] Discharge at the time of each measurement was taken to be that measured at the USGS gauge Provo River near Charleston. We also installed stage plates upstream and downstream of both bed load sampling sites. We measured the water surface slope over a distance of approximately 5 channel widths at the two sampling sites during each measurement. When flows receded after the flood, we conducted bed material point counts of the submerged bed in the vicinity of each sampling site to determine the grain size distribution of the bed surface.

3.1.2. Computation of Sediment Flux

[15] The bed load measurements from each site conform closely to a power function and showed no hysteresis. Transport rates were characterized using sediment rating curves,

$$Q_s = aQ^b, \tag{2}$$

where Q_s is sediment flux and Q is discharge. For each site we calculated cumulative sediment transport for the 2009 flood using mean daily discharge data provided by the USGS gauging station.

[16] We used a bootstrap approach to calculate the uncertainty associated with our estimates of the annual sediment load. For each data set, we generated 1000 random samples with replacement from the transport data. We fit a rating curve to each random sample and used the function to calculate total sediment load over the flood hydrograph. From the 1000 samples we generated a distribution and calculated the median value and 95% confidence interval to estimate influx, efflux, and net storage. We used a bulk density of 1855 kg m^{-3} to convert sediment mass to volume [*Bunte and Abt*, 2001].

3.2. Determining Change in Storage

[17] We divided the study area into seven reaches for the purpose of calculating change in sediment storage (Figure 3). Reach boundaries were defined to provide consistent within-reach properties based on channel planform and measurement technique. Reaches 1 and 2 are confined by a levee along the right bank protecting a wastewater treatment plant and, as a result, are relatively straight. In these reaches the channel is steep and there is little space in the channel to accommodate new deposits. In reaches 3 through 7 the river is relatively unconstrained, and the channel was constructed with a meandering planform.

[18] Three reaches (1, 4, and 6), accounting for nearly one-third of the study area, were surveyed before the flood using total stations and rtkGPS systems. It was not possible to survey channel topography in the remaining four reaches prior to the flood. Preflood bathymetery for reaches 2, 3, 5, and 7 was determined from a combination of aerial LiDAR and multispectral aerial imagery. We surveyed the entire study area after the flood using total station and rtkGPS surveys. We used the topographic data to construct pre- and postflood digital elevation models (DEMs). We computed changes in bed material storage using geomorphic change detection techniques [*Milan et al.*, 2011], which we describe below.

3.2.1. Direct Measurement of Topography via Ground Surveys

[19] In reaches 1, 4, and 6 we surveyed preflood topography during low flows in September–October 2008 and

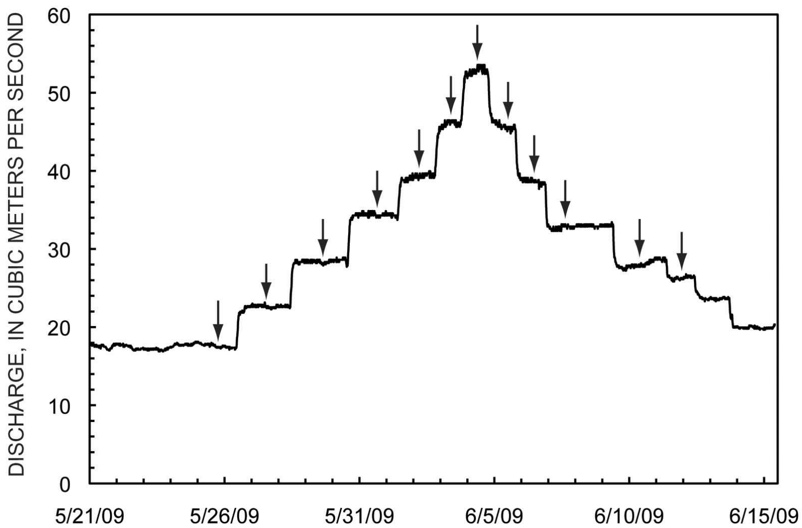

Figure 4. The 2009 flood hydrograph measured at USGS gauge 10155500, Provo River near Charleston. We selected the stair-step pattern to facilitate measurement of bed load transport rates. The flow was increased or decreased by approximately 5.7 $m^3 s^{-1}$ increments, and was held constant for 1–2 days at each discharge. Arrows indicate days when we sampled bed load transport.

postflood topography during October–November 2009. All survey data were collected in WGS84, using a Topcon HiperPro rtkGPS, Leica GS15 rtkGPS, and Leica TCRA 1203 + total station. The total station was used in portions of the channel that were too deep to safely survey with rtkGPS. Average point density of pre- and postflood surveys was approximately 0.32 points m^{-2}. Point densities were greater in areas with steeper or more complex topography and less dense in parts of the channel with little relief [*McCullagh*, 1981; *Brasington et al.*, 2000; *Valle and Pasternack*, 2006].

3.2.2. Remotely Sensed Topography

[20] In reaches 2, 3, 5, and 7 we mapped 2008 topography using a combination of remotely sensed data: LiDAR and multispectral (RGB) imagery. These data were acquired by Sanborn Mapping Inc. on 23 September 2008. We used the data to create a composite terrain model for each reach, using LiDAR for above-water locations and estimating submerged elevations by subtracting estimated channel depth from the water surface elevation [*Legleiter*, 2012]. Channel depth was estimated using a statistical relation between flow depth and spectral intensity of the RGB imagery [e.g., *Winterbottom and Gilvear*, 1997; *Marcus and Fonstad*, 2008; *Legleiter et al.*, 2009]. Although this approach is inherently less accurate than ground-based surveys, it provides useful information in the absence of other data. The Provo River provided ideal conditions to apply this technique; the flow is relatively shallow with very low turbidity, aquatic vegetation is minimal, and there is little overhanging riparian vegetation.

[21] We calibrated relations between RGB intensity and measured flow depth using the preflood survey data in reach 4. We collected these ground survey data within two weeks of the time when the air photos were acquired, and discharges during this time varied little (Table 1). We selected reach 4 because the ground survey was conducted within two weeks of aerial photographs and discharge and stage were nearly identical on both dates.

Table 1. Mean Daily Discharge and Stage for USGS Gauge 10155500, Provo River Near Charleston for Days With Aerial Photography or Ground Surveys[a]

	Date	Q ($m^3 s^{-1}$)	Stage (m)	Difference in Stage (m)
Aerial Photography Flight	9/23/2008	6.43	1.173	–
Ground Surveys				
Longitudinal Profile	10/12/2008	6.60	1.181	−0.008
Reach 1	9/12/2008	7.65	1.227	−0.054
Reach 1	9/17/2008	6.12	1.158	0.015
Reach 4	10/6/2008	6.31	1.167	0.005
Reach 4	10/7/2008	6.23	1.163	0.009
Reach 6	9/26/2008	6.09	1.156	0.016
Reach 6	9/27/2008	6.03	1.155	0.019
Reach 6	9/28/2008	6.06	1.155	0.018
Reach 6	10/3/2008	5.58	1.131	0.042
Reach 6	10/4/2008	5.86	1.145	0.028
Reach 6	10/5/2008	6.31	1.167	0.005

[a]The absolute difference in the stage on the date of the air photo flight and the date of the ground surveys ranges from 0.5 to 5.4 cm.

[22] For reach 4 we developed a multiple linear regression between depth and reflectance intensity of the three bands [*Winterbottom and Gilvear*, 1997]:

$$h = -0.08672 - 0.000374R + 0.000311G + 0.0000777B, \quad (3)$$

where h is depth of the water column, R is the reflectance of the red band, G of the green band, and B of the blue band. We evaluated a variety of relations between reflectance and depth, including ratios of bands [*Legleiter and Roberts*, 2005] and found equation (3) to be the best predictor of depth ($R^2 = 0.94$; Figure 5). We validated the relation using survey data from reaches 1 and 6, and found that the relations were a good predictor of depths in these reaches as well ($R^2 = 0.89$ and 0.92, respectively).

[23] We surveyed a longitudinal profile of water surface elevation along the channel centerline on 12 October 2008, and used these data to convert estimates of depth to absolute elevations. Discharges recorded at gauge 10155500 for the day of the air photo flight and the day of the water surface survey were 6.43 and 6.60 $m^3 s^{-1}$, respectively. This difference in flow corresponds to a difference in stage of 8 mm at the Charleston gauge. We neglected this stage difference because it is small relative to the magnitude of the uncertainty inherent in computation of depths using the RGB imagery (on the order of 20 cm). GPS points on the longitudinal profile were collected every 5–10 m, with an effort to survey points at locations where there was a change in water surface slope. We linearly interpolated the water surface between survey points to develop a continuous water surface profile. We then subtracted the estimated depths from the interpolated water surface profile to compute elevations. We merged the spectrally based bathymetry with the bare-earth topography generated from the LiDAR to create a composite DEM.

3.2.3. Computing Change in Storage

[24] With the hybrid mix of topographic survey data described above, we derived 1 m DEMs for the seven reaches for preflood and postflood conditions based on either ground survey or LiDAR and spectrally derived bathymetry. We selected a 1 m resolution because it adequately represents the topography of mapped geomorphic units and was supported by the point density available. We calculated the difference between DEMs for the pre- and postflood periods on a cell-by-cell basis to calculate a DEM of difference (DoD) using the geomorphic change detection software [*Wheaton et al.*, 2010]. The change detection software was used to (i) independently estimate the errors in the input DEMs; (ii) propagate those errors into the DoD change calculation; (iii) estimate the probability that calculated DoD changes are real; and (iv) use the probability estimates to exclude areas of change that were not above a selected confidence interval from the volumetric estimates of erosion and deposition.

[25] We used two techniques to estimate errors in the individual DEMs. In reaches 1, 4, and 6, where ground survey data were available before and after the flood, we used a spatially variable fuzzy inference system calibrated to rtkGPS and total station surveys to estimate DEM errors [*Wheaton et al.*, 2010]. The fuzzy inference system is based on the idea that construction of a DEM from survey data is a tradeoff between sampling intensity and the topographic complexity of the surface being surveyed. In reaches 2, 3, 5, and 7, where the spectral bathymetry technique was used to generate preflood DEMs, we used a more conservative spatially uniform estimate of DEM error. We assigned a 20 cm error to the preflood DEMs derived from the multispectral imagery (20 cm is the standard deviation of observed minus predicted elevations for reach 4) and a 6 cm uniform error for the post flood surveys.

[26] We calculated the combined error for the individual DEMs on a cell-by-cell basis using

$$E = \sqrt{e^2_{\mathrm{DEM1}} + e^2_{\mathrm{DEM2}}}, \quad (4)$$

where E is the combined, or propagated error, and e_{DEM1} and e_{DEM2} are the errors associated with the 2008 and 2009 DEMs [*Brasington et al.*, 2003]. We compared the propagated errors to the DoD to calculate a T-Score and estimate a probability that the calculated change was real, as described by *Lane et al.* [2003]. We used a more conservative 95% confidence interval in reaches 2, 3, 5, and 7, where we were less confident in the topographic surfaces generated from the multispectral imagery. In reaches where topography was directly measured with rtkGPS both pre- and postflood, we used a less conservative 80% confidence interval.

[27] We calculated change in storage by comparing net volume differences between erosion and deposition (i.e., deposition minus erosion). We calculated volumetric errors (±volume) for the estimates of scour and fill volumes by multiplying the estimated propagated DEM error on a cell-by-cell basis by the area of the cell. Those individual volumetric errors were used to estimate the total uncertainty in the net volumetric change in storage calculated for each reach. Additionally, to facilitate comparison of the magnitudes of change in each reach, we calculated the relative change in storage (a volume to surface area ratio) by dividing the net volumetric change by the total area of the reach.

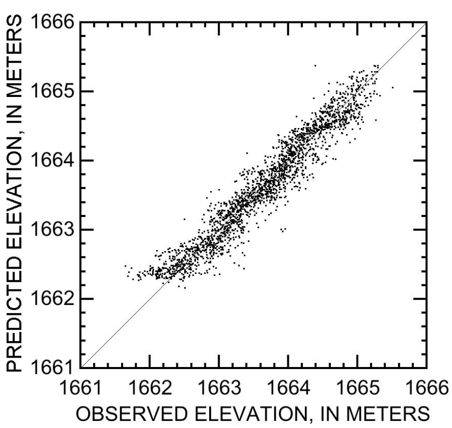

Figure 5. Relation between measured bed elevations and elevations derived from spectrally based bathymetry. Data is from reach 4, the reach used to develop the multivariate regression (equation (3)). The multivariate regression was used to model flow depths. Bed elevations were obtained by subtracting the spectrally derived depths from the water surface profile.

4. Results

4.1. Bed Load Flux

[28] Bed load measurements were made on the rising and receding limbs of the flood at both sites. Sampling began at the onset of detectable gravel transport. We collected 32 samples at Midway and 31 samples at Charleston. Measured transport rates ranged from 0.01 to 57 g m^{-1} s^{-1} at Midway and 0.2 to 27 g m^{-1} s^{-1} at Charleston (Figure 6). At both sampling sites, measured bed load transport rates showed a strongly nonlinear relation with discharge and did not display any hysteresis. Table 2 reports the values of parameters *a* and *b* from equation (2) that were determined for at each measurement station.

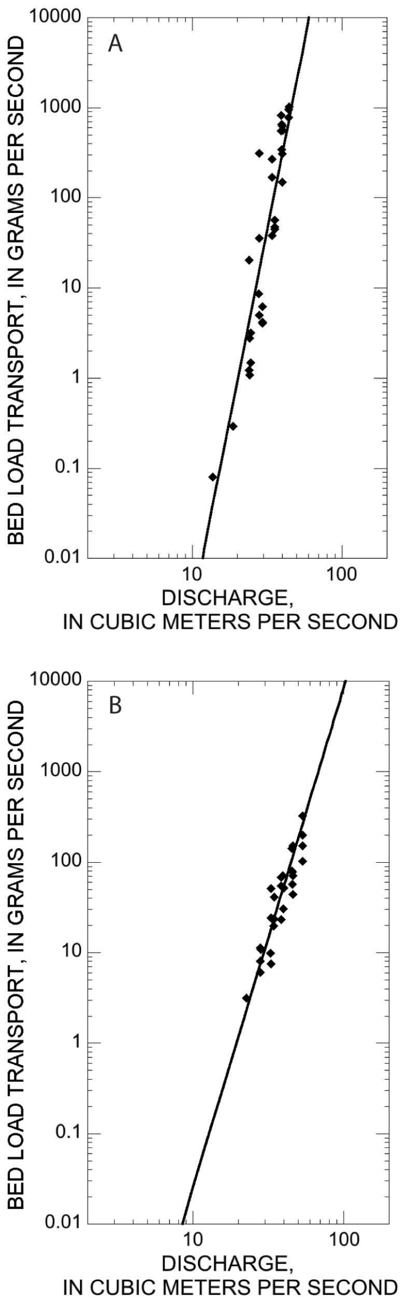

Figure 6. Bed load rates at (a) Midway and (b) Charleston. Diamonds indicate field measurements of bed load transport collected during the flood in 2009. The lines represent the sediment rating curves.

Table 2. Values of Coefficient *a* and Exponent *b* From Equation (2) That Were Determined for the Two Bed Load Measurement Sites, Midway and Charleston[a]

Sampling Site	*a*	*b*
Midway	1.863E-06	4.642
Charleston	2.865E-09	7.000

[a]In equation (2), Q is in units of m^3 s^{-1} and Qs is expressed in g s^{-1}.

[29] Total sediment loads computed from the sediment rating curves demonstrate that there was net accumulation during 2009. Approximately 5.84×10^5 kg (95% CI is 3.93×10^5 and 9.64×10^5 kg) entered at Midway and 5.96×10^4 kg (5.18×10^4 and 6.79×10^4 kg) exited at Charleston. These estimates correspond to sediment volumes of 319 m^3 (212–519 m^3) and 32 m^3 (28–37 m^3) at Midway and Charleston, respectively. Thus, despite the uncertainty associated with these estimates, these calculations demonstrate that bed load influx exceeded bed load efflux by an order of magnitude. The estimated net sediment accumulation based on these transport measurements is 287 m^3 (180–489 m^3).

4.2. Change in Storage

[30] Both scour and fill were small in reaches 1 and 2, where the channel is confined along the right bank by a levee (Table 3; Figures 7 and 8). Both scour and fill increased in reaches 3 and 4, where the channel is more able to adjust. Scour and fill were both larger in reach 4 than in any other reach. Deposition remained large in reaches 5, 6, and 7 and deposition exceeded erosion in all reaches except 1 and 3. The only reaches in which the net storage exceeded the uncertainty in erosion and deposition estimates were reaches 1 and 5 (Table 3). If minimum and maximum error values are propagated in the along stream accumulation of scour, fill, and storage, the net reach storage of 472 m^3 is dwarfed by the accumulated error of $\pm$1344 m^3 (Figure 8). This error estimate is likely too large, however, because it assumes that the minimum and maximum errors consistently propagate from one reach to the next. In contrast to the net storage, both scour and fill estimates exceed the error bounds in all reaches except for erosion in reach 6. The cumulative scour and fill for the entire study reach is roughly twice that of the cumulative error in each term (Table 3), even using the simple and conservative accumulation of minimum and maximum errors. Both cumulative scour (1454 m^3) and fill (1926 m^3) are significantly larger than the measured sediment flux at the upstream (319 m^3) and downstream (32 m^3) ends of the study reach.

[31] The calculated change in storage computed from the analysis of pre- and postflood topography is consistent with the imbalance in bed load transport described above. Based on the volumes of sediment scour and fill calculated from DoDs developed for each reach (Figure 7), we computed approximately 472 m^3 ($\pm$1344 m^3) of net sediment deposition, resulting from approximately 1454 m^3 ($\pm$796 m^3) of scour and 1926 m^3 ($\pm$1043 m^3) of fill (Table 3).

Table 3. Calculated Change in Storage for All Seven Reaches[a]

Reach	Scour (m^3)	Fill (m^3)	Net Storage (m^3)	Cumulative Storage (m^3)	Reach Area (m^2)	Relative Change in Storage (m^3 m^{-2})
1	122 (±22)	51 (±43)	−71 (±49)	−71 (±49)	4376	−0.016
2[b]*	172 (±51)	208 (±87)	36 (±100)	−35 (±149)	10,170	0.004
3*	303 (±200)	283 (±126)	−20 (±237)	−55 (±386)	19,573	−0.001
4	311 (±190)	440 (±293)	129 (±349)	74 (±735)	10,972	0.012
5*	106 (±30)	306 (±135)	200 (±139)	274 (±874)	16,752	0.012
6	183 (±223)	348 (±255)	165 (±339)	439 (±1213)	10,692	0.015
7*	257 (±80)	290 (±104)	33 (±131)	472 (±1344)	13,514	0.002
1 + 4 + 6	616 (±435)	839 (±591)	223 (±737)	NA	26,040	0.009
Σ1–7	1454 (±796)	1926 (±1043)	472 (±1344)	NA	86,059	0.005

[a]Scour, fill, net change in storage, and cumulative downstream change in storage were determined from the DEMs of difference (DoDs). Relative change in storage is net volumetric change divided by reach area.
[b]Asterisks denote preflood topography determined using surveyed water surface elevation and water depth estimated from aerial photography.

[32] The morphologic changes documented during the flood were subtle relative to the uncertainty associated with the topographic measurements. Figure 9 depicts two different DoDs for reach 4 that were calculated using the same topographic inputs. One DoD—the gross DoD—does not consider uncertainty (Figure 9a). The other DoD incorporates the spatially variable uncertainty threshold (Figure 9b) that was used to calculate changes in storage in this study. The computed volumes of scour, fill, and net channel change differ substantially between the two DoDs, highlighting the significant effect of uncertainty on the budget calculations. The DoDs shown in Figures 9a and 9b correspond to histograms (Figures 9c and 9d, respectively), illustrating the large proportion of the channel where the magnitude of the topographic change did not exceed the defined uncertainty threshold. Areas not included in the budget

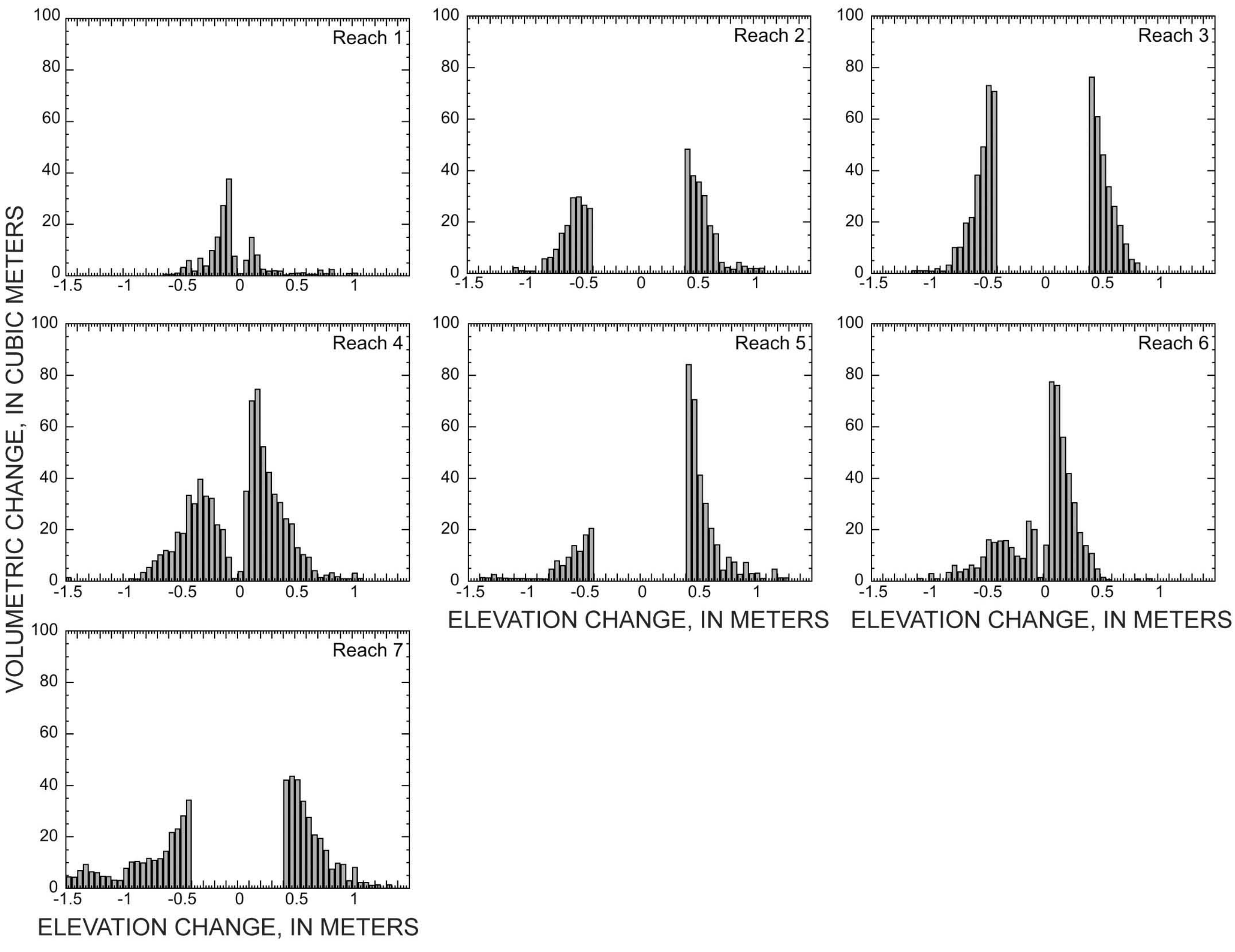

Figure 7. Volumetric change in storage for each reach calculated from the DEMs of difference (DoDs).

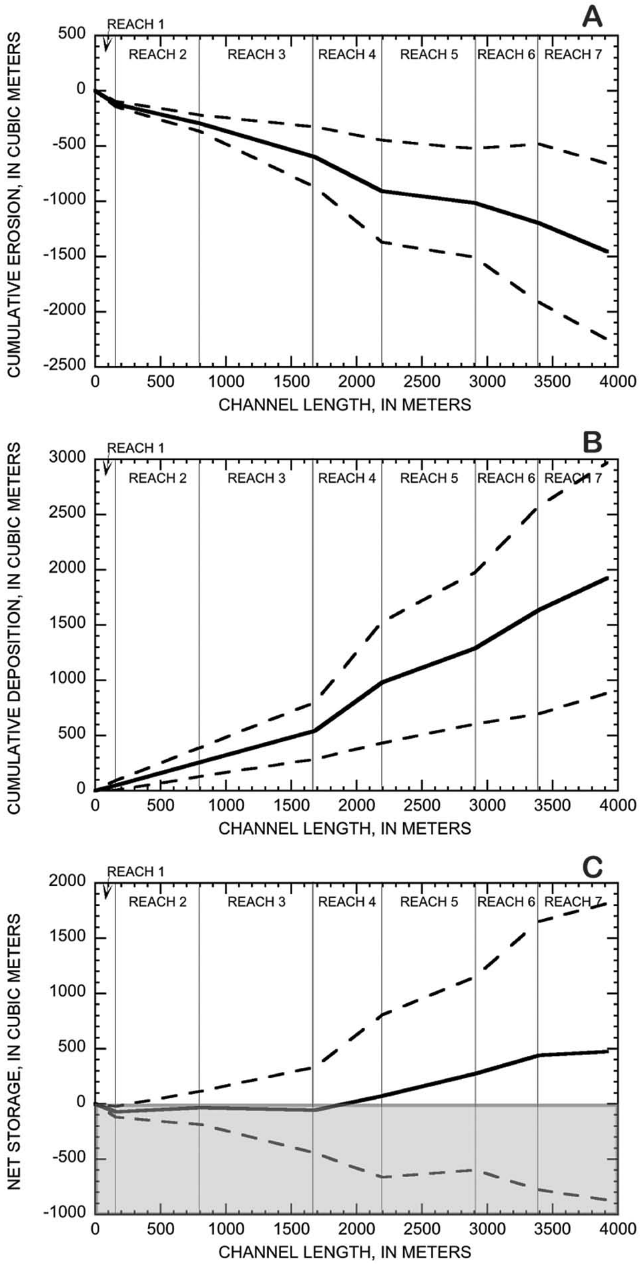

Figure 8. Cumulative (a) erosion, (b) deposition, and (c) net sediment storage in the study area, based on the analysis of the topographic data. The dashed lines represent the cumulative uncertainty.

calculations are primarily those with small elevation change. The impact of the uncertainty analysis on our estimates of change in storage is apparent in the histograms of channel change derived for the DoDs from all seven reaches (Figure 7). Distributions from reaches 2, 3, 5, and 7 are truncated because small changes were excluded from the analysis due to the more conservative uncertainty threshold applied to the DEMs derived from LiDAR and multispectral imagery.

[33] The error bars bracketing our estimated net change in storage are relatively large (± 1344 m^3; Figures 8 and 10) because the propagated error estimates are large relative to the modest magnitudes of net change that actually occurred (Table 3). It should be noted that even though the estimated DEM errors are much larger in the reaches where preflood topography was derived from multispectral imagery (reaches 2, 3, 5, and 7), these errors do not necessarily result in larger estimates of volumetric uncertainty (Table 3, column 4). This is because the volumetric error is based only on the areas in the DoDs where the magnitude of channel change exceeded the calculated uncertainty threshold.

4.3. Closure of the Sediment Budget

[34] With a flux estimate of approximately 290 m^3 of net aggradation and an estimate of change in storage of approximately 470 m^3 of net aggradation, both the change in storage as determined from bed load transport and from topographic data demonstrate that sediment accumulation occurred during the 2009 flood (Figure 10). Had the analysis relied solely on morphologic data, however, the budget would have been indeterminate. The magnitude of the uncertainty for reaches in which pre- and postflood topography were measured via rtkGPS (reaches 1, 4, and 6; Table 3, column 4) suggests that this would have been the case even if we were to have surveyed the entire study area prior to the flood using traditional ground survey techniques.

5. Discussion

5.1. Provo River Sediment Budget

[35] There are inevitable uncertainties associated with calculation of each term of a sediment budget. Had we solely based the budget on topographic measurements, we would not have reached a conclusion consistent with earlier observations of aggradation. However, estimates of change in storage calculated from flux measurements of input minus export unambiguously demonstrate that aggradation occurred during the 2009 flood, because the volume of accumulation greatly exceeded the uncertainty in the transport measurements. Had we only estimated change in storage from topographic measurements, we would not have reached this conclusion because the topographic changes that occurred were small in relation to the uncertainty in those measurements.

[36] *Grams and Schmidt* [2005] emphasized that without faithful accounting of uncertainty a budget may be considered closed, when in fact, it is indeterminate. Our findings provide a reminder that wherever the uncertainty exceeds the absolute value of the net change in sediment storage, a budget is unavoidably indeterminate. Thus, it is critical to use the appropriate technique at the appropriate temporal and spatial scale to have the best chance of avoiding indeterminacy in sediment budgets.

[37] In the case of the 2009 flood season budget for the Provo River, the topographic changes were subtle and the uncertainty was large. The theoretical basis for and application of multispectral and hyperspectral imagery to quantify channel depths is well documented using both empirical and theoretical approaches [*Winterbottom and Gilvear*, 1997; *Wright et al.*, 2000; *Whited et al.*, 2002; *Legleiter et al.*, 2009], and the technique performs well in systems where the water is relatively clear, shallow, and free of aquatic vegetation, such as in our study area. Despite having near ideal conditions for application of this technique, the DEMs derived from spectral bathymetry provided less accurate representations of channel topography than the

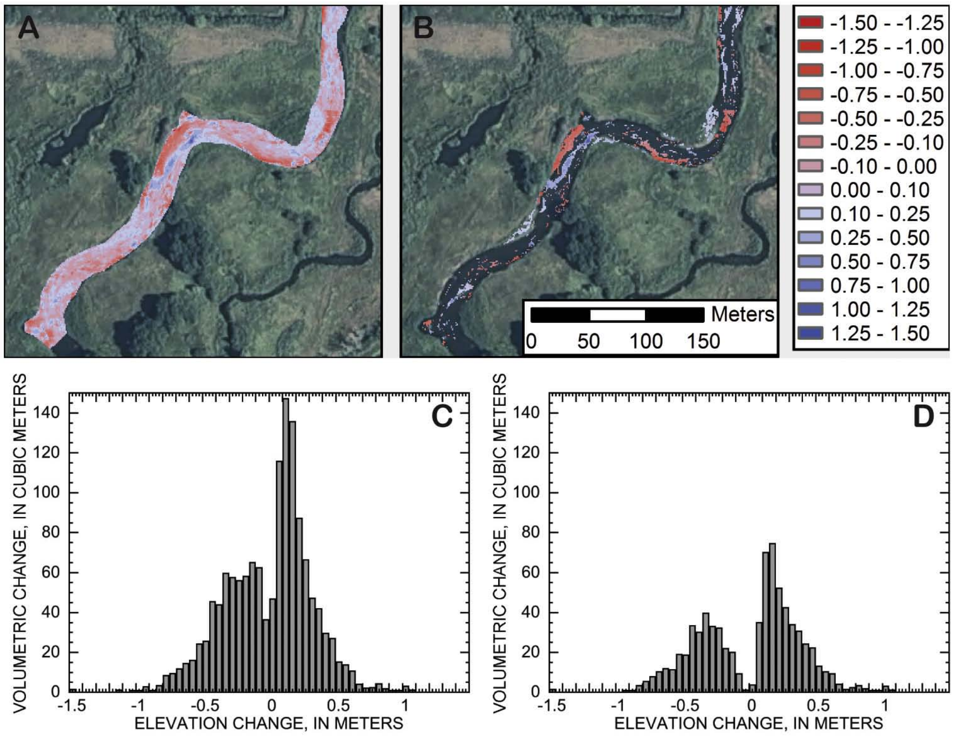

Figure 9. Volumetric change in storage for reach 4 computed from (a) the gross DoD and (b) the DoD calculated using the spatially variable approach for quantifying uncertainty, as described in the methods. (c) A histogram computed from the gross DoD, corresponding to Figure 9a; (d) a histogram computed from the DoD calculated using the spatially variable uncertainty threshold (Figure 9b).

DEMs derived from high-density ground surveys and necessitated a large threshold of detection in reaches 2, 3, 5, and 7. Yet even in reaches where both the pre- and post-flood DEMs were derived from ground survey data, the thickness of deposition and erosion in large areas was less than the inherent uncertainty of those measurements (Figure 9).

5.2. Net Change, Total Change, and Channel Activity

[38] Despite the large uncertainty associated with topographic differencing, these measurements provided unique insights into system behavior. Although the flux measurements provided a better-constrained estimate of large-scale change in storage, direct measurements of topographic

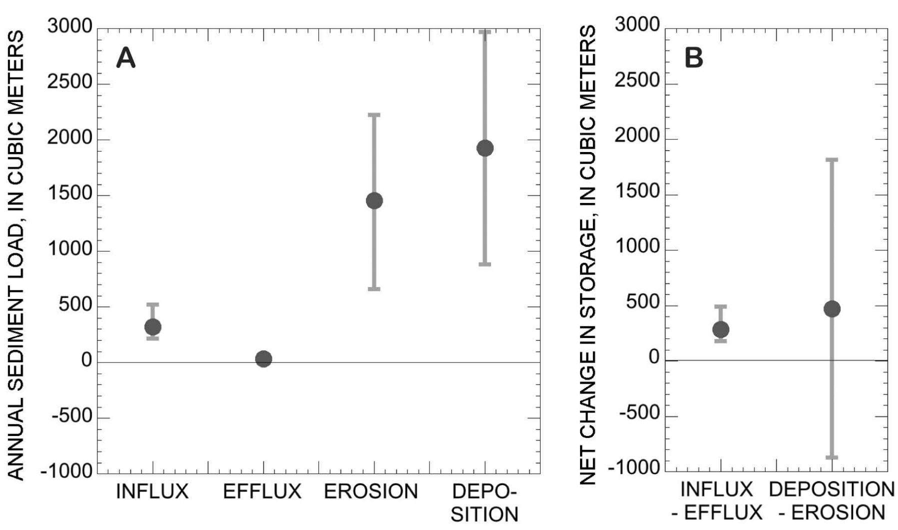

Figure 10. The sediment budget for the 2009 flood. (a) The four measured components of the budget: estimated influx and efflux (calculated from the sediment rating curves), and deposition and erosion (determined from topographic measurements). (b) The change in storage computed from each of these components of the budget.

change provide documentation of spatial patterns of morphologic adjustment. Additionally, the topographic data provided an estimate of the relation between net channel change and total channel change (or channel activity). For this particular scale of study area and study period, the magnitude of total scour and fill was significantly larger than the magnitude of the net difference between influx and efflux (Figure 10). Total channel change—calculated as the sum of the total volume of scour and of fill—greatly exceeded the magnitude of the net change, indicating that there was significant local reorganization of sediment within the study area.

[39] Even in this "event-based" study there were inconsistencies between the time domain of the flux measurements and the time when change in storage was measured. Most budgets are calculated over larger areas and over longer time periods. Morphological sediment budgets, for example, are often integrated over many bed-mobilizing floods. We sought to minimize these inconsistencies by calculating a sediment mass balance for a discrete flood event. Topographic data were collected during the course of a year and sediment flux data were collected during a 3 week period, but the dam-controlled hydrology limited bed-mobilizing flows to a single period between the topographic measurements. The controlled dam release presented a relatively unique field opportunity to isolate the erosion and deposition associated with a discrete flood; when integrating over multiple bed-mobilizing flows, the occurrence of compensating scour and fill make it probable that measurements of channel change are underestimates of total channel activity [*Lindsay and Ashmore*, 2002].

[40] For the Provo River sediment budget, we suspect that more accurate measurements would likely amplify the difference between the total volumetric change in storage and the total volume of sediment flux. With higher resolution topographic measurements, smaller topographic changes would have been detectable. This would have improved our ability to document both total and net channel change. Because compensating scour and fill cannot be measured from before and after topography, measured volumetric change will inevitably be an underestimate of total volumetric change. These factors suggest that the magnitude of the erosion and deposition terms may be even larger relative to the net change in storage in the Provo River sediment budget. Although the reach studied here had been rebuilt 5 years before the dam release, such that extensive reworking is not surprising, our work nonetheless is a reminder that an inference of total channel activity from net sediment flux should be verified by observations of channel change within the reach.

5.3. Uncertainties and Implications for Fluvial Sediment Budgeting

[41] Development of sediment budgets is an essential exercise in geomorphology, used across the discipline in theoretical and applied studies. This study provides some general insights into the challenges and uncertainties associated with developing a reach-scale budget in a fluvial setting. Closing a sediment budget—matching measured changes in storage with calculated differences between inputs and outputs—is a difficult task even in well-constrained settings. Although we attempted to quantify all components of the sediment budget for the Provo River, in some cases the uncertainties associated with different budget terms exceeded the measured value.

[42] Quantification of inputs and outputs is inevitably subject to the uncertainty associated with estimating transport rates. The magnitude of uncertainty that is associated with our estimates of sediment flux is a reflection of the fact that bed load transport rates exhibit great spatial and temporal variability [*Ashmore and Church*, 1998; *Hicks and Gomez*, 2003]. Even under steady flow conditions, bed load transport rates are highly variable [*Davies*, 1987; *Gomez*, 1991]. Bed load transport is also difficult to measure. The bed load transport predictions we developed for the Provo River represented a significant effort to constrain the estimates of influx and efflux by directly measuring transport rates during a controlled flood, in a system without any supply limitation or hysteresis. Despite these efforts and despite the unique sampling opportunity, there is still unavoidable imprecision in our estimates of sediment flux. However, for the temporal and spatial scale of the budget developed here, the uncertainty associated with our estimates of change in storage based on flux measurement was much less than the uncertainty based on measurements of topographic change (Figure 10). Thus, for a relatively simple system such as the Provo River, direct measurements of flux may provide a better estimate of net change in storage.

[43] The two computational approaches to calculating a sediment budget—the difference in flux measurements and the difference between topographic surfaces—have different merits and different limitations that depend on the temporal and spatial scales of analysis and depend on the way that uncertainty propagates through time and space. For example, over shorter time steps, morphologic adjustment may be widely distributed and small relative to grain size of the bed material. Thus, the magnitude of erosion and deposition may be measureable but the difference between the two, the net storage, may not be detectable. A standard approach to dealing with uncertainty when determining differences between two DEMs is to establish a minimum level of detection threshold, below which change in elevation is neglected [*Brasington et al.*, 2000]. The approach used here *Wheaton et al.* [2010] is an attempt to improve upon this standard approach by incorporating knowledge of data quality, density, and topographic complexity to refine estimates of uncertainty. Nevertheless, this approach does nothing to constrain or limit uncertainties, but is simply one approach for determining whether the "signal" of topographic change exceeds the associated "noise" [*Milan et al.*, 2011]. In systems where change is subtle and there is substantial uncertainty associated with the strategy used to measure topography, it may not be possible to accurately identify real morphologic adjustments.

[44] When channel change is progressive toward sediment accumulation or evacuation, such as on the Provo River, a morphological budget may be better equipped to detect changes in storage when the budget is computed over a longer time domain. The effects of time and error propagation on budget calculations are conceptually represented in Figure 11. In Figures 11a and 11b, t_1 represents the budget calculated for the 2009 flood. T_2 and t_3 represent two hypothetical floods of longer duration that were generated using hydrologic data for previous flood releases from Jordanelle Dam. When there is small topographic change, as in 2009, the uncertainty associated with estimates of topographic change is proportionally a large part of the net

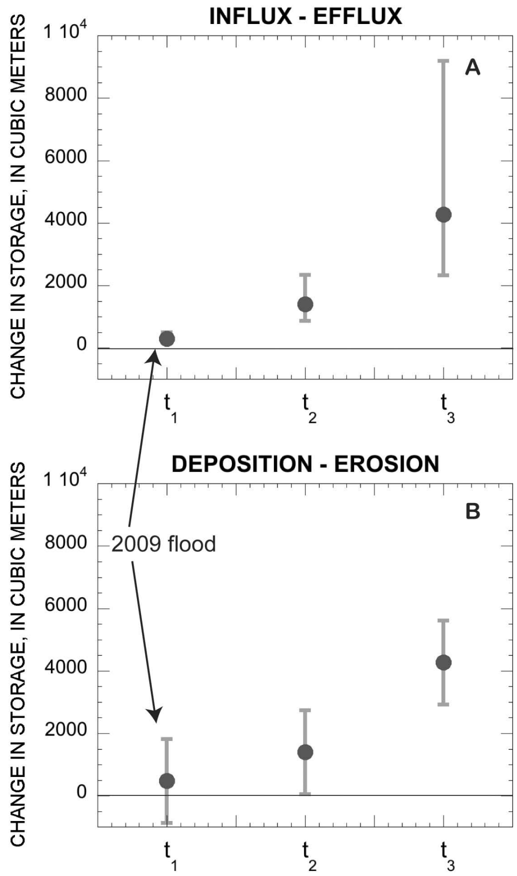

Figure 11. A conceptual plot depicting the effects of time and error propagation on sediment budgeting. T_1 shows the estimated change in storage for the 2009 flood as calculated from (a) the difference between sediment inputs and outputs and (b) the difference between topographic surfaces. The uncertainty associated with measurements of flux is additive and propagates through time, whereas the uncertainty associated with calculation of ΔS based on topographic data remains relatively constant as the time step increases.

change in storage. For the budget presented here, the uncertainty bars bracket zero change in sediment storage, demonstrating that the budget would be indeterminate if based on changes in topography alone. In 2009 (t_1) the uncertainty associated with our estimates of sediment flux were proportionally very small as compared to the net change in storage. Thus, only the sediment flux measurements demonstrated that the system accumulated sediment during the 2009 flood.

[45] Assuming the sediment rating curves remain stable through time, the relations can be used to calculate net flux for t_2 and t_3 (Figure 11a). Although the uncertainty associated with our flux-based estimate of net storage was proportionally very small for the 2009 budget, the uncertainty associated with change in storage derived from measurements of sediment flux is additive and propagates through time. Therefore, at t_2 and t_3 the uncertainty represents a larger proportion of the net change in storage. In contrast, the uncertainty does not propagate through time when ΔS is determined from topographic measurements. In Figure 11b the point estimates of change in storage for t_2 and t_3 are assumed to be equal to those derived from the calculated fluxes. The uncertainty remains constant through time because calculation of uncertainty for the two topographic surfaces is independent of the time when the surface was measured. Thus, in a system that is progressively accumulating or evacuating sediment, as time progresses it becomes possible to develop a determinate budget based on morphologic data, whereas flux-based estimates will become less informative as the time domain of the budget increases. For example, if the net change in storage remained small at t_3 (e.g., 2000 m^3), the budget would only be determinate if informed by topographic data, and would be indeterminate if ΔS were calculated as the difference between sediment input and export.

6. Conclusions

[46] We measured all components of a sediment mass balance for a single, dam-controlled flood on a reconfigured gravel bed river. Flow was released from Jordanelle Dam on the Provo River in a fashion that allowed measurements of bed load transport during steady flow at sections above and below the study reach. Detailed topographic data were collected before and after the dam release. Based on transport rate measurements, sediment input to the reach exceeded outputs, producing a net sediment accumulation of approximately 290 m^3 (95% CI is 180–489). The difference between the erosion and deposition was also positive (472 m^3), but uncertainty in the topographic differencing ($\pm$1344 m^3) was larger than the observed net change in storage. Both the change in storage calculated from flux measurements and via topographic changes suggest that the system accumulated sediment during the 2009 flood. However, had the analysis relied solely on morphologic data the budget would have been indeterminate. Because the morphologic adjustments that occurred during the 2009 flood were relatively subtle, a morphologic sediment budget would have been indeterminate even if we had measured pre- and postflood topography for the entire study area using ground survey techniques.

[47] When developing sediment budgets for which the primary interest is to quantify net accumulation or evacuation of sediment, measuring sediment flux may be a better strategy if (1) appropriate sampling locations can be found, (2) the system is not subject to supply limitation or patterns of hysteresis, and (3) the time domain of the analysis is relatively short. However, because uncertainty associated with estimates of sediment flux propagate through time, over longer time frames, or in systems where topographic adjustments are pronounced, a morphological sediment budget may provide superior results. Direct measurements of topographic change is also necessary in order to accurately determine the total channel change, or channel activity, and to describe spatial patterns of morphologic adjustment within a reach.

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